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3430 Noriega
San Francisco, CA

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An Improved Radiative Transfer Model for Climate Calculations

Robert W. Bergstrom¹

Eli J. Mlawer²

Irina N. Sokolik³

Shepard A. Clough²

Owen B. Toon³

¹Bay Area Environmental Research Institute, 3430 Noriega, San Francisco, CA 94122

²Atmospheric and Environmental Research Inc., Cambridge, Mass

³Department of Oceanic and Atmospheric Sciences, University of Colorado, Boulder CO

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ABSTRACT

This paper presents a radiative transfer model that has been developed to accurately predict the atmospheric radiant flux in both the infrared and the solar spectrum with a minimum of computational effort. The model is designed to be included in numerical climate models

To assess the accuracy of the model, the results are compared to other more detailed models for several standard cases in the solar and thermal spectrum. As the thermal spectrum has been treated in other publications (Mlawer et al., 1997, Sokolik et al., 1998) we focus here on the solar part of the spectrum. We perform several example calculations focussing on the question of absorption of solar radiation by gases and aerosols.

1. Introduction

In general, the calculation of atmospheric radiation has been a challenging task. The methods range from very detailed line-by-line models with scattering (Moncet and Clough, 1997) that can take hours on the fastest computers for a single simulation to parameterizations used in global climate models that are simple empirical formulas (Barker and Li, 1995). Climate modelers have been faced with the problem of spending a large amount of computer time for detailed calculations or accepting approximate methods with limited accuracy. Not surprisingly, the Intercomparison of Radiation Codes in Climate Models (ICRCCM - Ellingson and Fouquart, 1991) project showed that there was a wide range in the results from different atmospheric radiation computer programs even for relatively simple cases. (For example, the case of gaseous absorption and Rayleigh scattering showed a range of 47 W/m² in the predicted solar flux at the surface (Fouquart et al., 1991).)

The problem for climate modelers has gotten worse with the increased emphasis on atmospheric aerosols. In the case of the atmospheric aerosol, the global effect on the atmospheric radiation may be on the order of 1 W/m² (IPCC, 1995) so that climate modelers would like to have tools that are in that range. At the same time, however, few modelers are willing to spend the majority of their calculation resources on the atmospheric radiation calculation.

To further complicate the situation, comparisons between measurements and calculations of the solar radiation at the surface have shown relatively consistent differences on the order of tens of W/m² even under clear skies (Wiscombe 1996, Kato et al., 1997). There is considerable debate about the causes of this discrepancy, ranging from experimental error, aerosol absorption, water vapor absorption and even an unknown gas absorption (Kato et al., 1997). The situation for the thermal spectrum is somewhat better in that comparisons between calculations and measurements can be routinely within 1 W/m² (Clough et al., 1996). In fact, the limiting factor in comparing the measurements and predictions in the thermal spectrum has been the measurement of the temperature profile (Clough et al., 1996).

In an attempt to maximize accuracy and minimize the amount of computational effort in the calculation of atmospheric radiation, we have developed a simplified tool to compute the radiative flux in both the solar and infrared. (A previous version of the model was presented at the 1996 IRS symposium (Bergstrom et al., 1996).) In this paper we briefly set out the formulation of the model and concentrate on estimates of the accuracy of the predictions. As the thermal spectrum has been discussed in other publications (Mlawer et al., 1997, Sokolik et al., 1998) we focus here on the solar part of the spectrum. We present some results for two different types of atmospheric aerosols and show that the

difference between the aerosol types is considerably larger than the uncertainty in the model due to the two stream method. While the model is applicable to clouds, we concentrate in this paper on aerosol properties.

2. Radiative Transfer Model

A radiative transfer model must do two things. First, specify the radiative properties of the atmospheric constituents and second, compute the vertical profile of the radiant fluxes (or intensities) from the radiative transfer equation.

2.1 Description of the Absorption and Emission by Atmospheric Gases – the k distribution method

One of the most difficult problems in radiative transfer has been combining the gaseous absorption of a vibrational-rotational band of a gas with the scattering of a cloud or aerosol. The main reason that the problem is difficult is that each band has literally thousands of separate lines. The spacing of the lines is so small that the spectral interval necessary to resolve each individual line is on the order of 0.01 cm⁻¹. The calculation of radiative transfer at this spectral resolution is called a line-by-line code (LBL). Such a resolution is impractical for a climate model (there are roughly 5 million 0.01 cm⁻¹ intervals in the atmospheric radiation spectrum). Traditionally, to save computational effort, band models have been used to compute the transmission over a spectral interval (Goody and Yung, 1989). However, combining the gas absorption with aerosol or cloud scattering was then approximate and the accuracy difficult to assess.

Recently, Mlawer et al., (1997) and others (most recently Goody et al., 1989; Lacis and Oinas, 1991; Fu and Liou, 1992; Kratz, 1995; Kato et al., 1997) have used the k distribution method to develop sets of k values for the thermal or solar spectrum. The basis of the method is that the transmission over a frequency interval, $\Delta\nu$, can be rewritten in terms of the cumulative probability variable, g. Thus, the values of the absorption coefficient in frequency space are mapped into the values of the absorption coefficient in cumulative probability space.

Mathematically, the mapping is to replace the average transmission, T, for a frequency interval (v_1, v_2) from

$$T(z) = \int_{v_1}^{v_2} \exp\left(-\int_{z_1}^{z_2} \kappa_v(p, \Theta) \rho dz\right) dv / (v_2 - v_1) \quad eq(1)$$

with the following expression:

$$T(z) = \int_0^1 \exp\left(-\int_{z_1}^{z_2} \kappa_g(p, \Theta) \rho dz\right) dg \quad eq(2)$$

where g is the cumulative probability, κ is the absorption coefficient, p is the pressure, Θ is the temperature, ρ is the density of the absorber, z is the height.

The reason for mapping the absorption coefficient into g space becomes clear if one plots the absorption coefficients at a fine spectral resolution then plots the same absorption coefficients as a function of the cumulative probability variable. (see Mlawer et al., 1997 – Fig. 1; Kratz, 1995 – Fig. 1; Fu and Liou, 1992 – Fig. 1; Lacis and Oinas, 1991 – Fig. 1). Another way to think of the k distribution method is that in a given frequency interval, you can calculate the radiative transfer for all identical k values at once instead of several times. Integrating the absorption coefficient in probability space simply takes fewer points than integrating the coefficients in frequency space.

Then, assuming the scattering properties are constant over the frequency interval, the single scattering albedo is simply the scattering coefficient divided by the sum of the scattering coefficient and the absorption coefficient for the interval. (The accuracy of assuming that the scattering properties are constant over the interval obviously depends on the size of the interval and the scattering and absorption properties of the aerosol or cloud.)

The strategy for developing the set of k tables for the thermal spectrum is set forth in Mlawer et al., (1997) and a similar strategy was adopted for the solar spectrum (Mlawer et al., 1998). In brief, the location and width of the bands were chosen to match the location of the major gas absorption bands. Each band can have two species with substantial absorption and a species that is responsible for substantial absorption is termed a “key species.” Other species in the band are referred to as “minor species.” The bands and species are listed in Table 1 for both the thermal and solar spectrum. The key

species are designated for two separate altitude regions, the transition being located near the tropopause for all bands.

Each spectral band is divided into 16 intervals in g-space chosen to have modified half Gauss-Legendre quadrature spacing. The boundaries and weights are shown in Mlawer et al., 1997- Table 2. The half Gauss-Legendre spacing is modified to place seven intervals between $g = 0.98$ and $g = 1.0$ in order to accurately determine the transmission for circumstances in which the transmission is dominated by the centers of the spectral lines in the band (such as low pressure).

The reference k_g values are stored for 59 pressure levels from 1050 to 0.01 mb in equal log pressure intervals. Log pressure is the variable used for interpolation. For each reference temperature level, the k_g 's are stored in +/-15K, +/-30K arrays where the reference temperature is the temperature corresponding to the pressure in the Midlatitude Summer profile. Using these tabulated absorption coefficients, the k_g 's for any arbitrary layer can be calculated by interpolation.

Overlap of gases (such as H₂O and CO₂) is calculated by using an additional variable that is related to the abundance of the two absorbing gases. The k_g 's for the combination of gases are stored for reference ratios so that the value for an arbitrary ratio can be calculated. Thus, the absorption coefficient tables have three variables: pressure, temperature and (when needed) the absorbing gas ratio.

In the thermal spectrum, the spectral variation of the Planck function is accounted for by determining an average value of the Planck function for all frequencies in a subinterval of the band (see Mlawer et al., 1997 – Eq. 11). The same approach is used to account for the spectral dependence within a solar band of the solar source function and the Rayleigh scattering coefficient. Thus, the solar source function and the Rayleigh scattering coefficients are not treated as constant across the band. The solar spectrum was taken from Kurucz (1995).

2.2 Description of the Scattering, Absorption and Emission by Aerosols

The radiative properties of atmospheric aerosols have been studied for quite some time (see the references in Bohren and Huffman, 1989; D'Alimeda et al., 1991; White, 1986; Horvath, 1993). Unfortunately, the properties are highly variable – both in time and space. When they are not measured or computed from a detailed aerosol model, they are often assumed or calculated from “representative” constituents (sulfates, soot, mineral aerosols, etc.). These representative models, such as the WMO or AFGL models (D'Alimeda et al., 1991, Shettle and Fenn, 1989) have serious limitations; in particular they imply that the aerosol is less variable than it actually is. For example, black carbon or soot does not

have a unique set of refractive indices and the specific absorption coefficient can vary significantly (Horvath, 1993; Colbeck et al., 1989). Similarly, mineral aerosol from the same desert region may have significantly different radiative properties (Sokolik et al., 1998).

Since we use a two-stream approach (discussed below), the aerosol radiative properties needed are the optical depth, single scattering albedo and the asymmetry parameter for each layer and spectral interval. The range of values for these quantities for some atmospheric aerosols is shown in Table 3 for both the solar spectrum and the thermal spectrum. As shown in Table 3, atmospheric aerosols (particularly mineral aerosols) have a large range of radiative property values.

2.3 Calculation of the fluxes from the equation of transfer - the two stream approximation

The radiative transfer equation can be written in terms of the intensity I at a particular frequency ν as

$$\begin{aligned} \mu \partial I_\nu / \partial \tau = & -I_\nu + \omega / 4\pi \int_{4\pi} P(\Omega_1 \rightarrow \Omega_2) I_\nu(\Omega_1) d\Omega_1 \\ & + S_\nu(\tau, \mu, \phi) \end{aligned} \quad eq(3)$$

where $\omega = \sigma/(\sigma+\kappa)$ is the single scattering albedo, σ is the scattering coefficient, κ is the absorption coefficient, P is the scattering distribution function, μ is cosine of the zenith angle, Ω is the solid angle, $\tau = \int (\kappa + \sigma) dz$ is the optical depth and z is the height.

For an emitting atmosphere

$$S_\nu = (1 - \omega) B_\nu(T)$$

where $B(T)$ is the Planck function at temperature T . While for a purely external source at a solar wavelength

$$S_\nu = \omega_\nu / 4 F_{sv} P_\nu(\mu, -\mu o, \phi, -\phi o) \exp(-\tau / \mu o)$$

where πF_{sv} is the solar beam and μ_0, ϕ_0 is the incident direction of the solar beam (μ is the cosine of the zenith angle and ϕ is the azimuthal angle. The use of the expression for the solar source implies that an additional direct component of the radiation field is present which must be added to the intensity to compute the total radiation field.

For this model we use the generalized two-stream method (Toon et al., 1989) to compute the radiant flux for an absorbing, emitting, and scattering inhomogeneous medium. The two-stream coefficients are selected to be the delta-quadrature for the solar spectrum and the hemispheric mean for the thermal spectrum. The coefficients depend only on ω, g and μ_0 (see Toon et al., (1989) - Table 1). The solution to equation (3) is the straightforward and for inhomogeneous layers results in a tri-diagonal matrix that can be easily inverted.

We should note that the solution to the equation of transfer (here a multi-stream approach) is completely separate from the k distribution method. The model discussed here is designed to be used in climate models where speed is at a premium, while a model used to analyze data would probably require more streams, such as a delta four stream approach (Fu and Liou, 1992) or perhaps even 16 streams. (For example, the AER code RRTM uses DISORT with 16 streams to analyze solar radiation data at the surface, (Mlawer et al., 1998).) However, as we discuss below, the current uncertainty in the aerosol radiative properties is larger than the error caused by the used of a two-stream approach. Therefore, we are using the two-stream code for our climate models.

3. Validations

The accuracy of a given model is always difficult to quantify precisely. While we present comparisons of our model results with more detailed calculations, there remains an obvious need for more benchmark calculations. We have used ICRCCM test cases (Fouquart et al., 1991) so that there is some consistency in our comparisons. As in the ICRCCM study, we focus here only on the downward radiant flux at the surface and the upward radiant flux at the top of the atmosphere.

3.1: Case 1- Gaseous Absorption and Emission

The solution of the equation of transfer for gaseous absorption and emission is straightforward for both the solar and thermal spectrum. These calculations can be compared to line by line calculations.

3.1.1 Solar Spectrum

We present the results for the solar transmitted beam at the surface from our model and the results for the line by line code LBLRTM (Clough et al., 1992; Clough and Iacono, 1995) in Table 5 for the Midlatitude Summer Atmosphere (MLS) and solar zenith angle of 0° .

The downward solar flux at the top of the atmosphere (TOA) is also shown. Table 5 indicates that the difference between our model and the line by line calculation is about 1 W/m^2 or less than 0.1%. The total absorption by the gases is about 244 W/m^2 .

3.1.2 Thermal Spectrum

Mlawer et al., (1997) show that the correlated k method can compute fluxes in the thermal spectrum to 1 W/m^2 accuracy and heating rates to 1% accuracy. This is well within the criterion for a climate model. The results for our model are essentially identical to Mlawer et al., (1997) so they need not be presented here.

3.2 Case 2 - Rayleigh Scattering and Gaseous Absorption – Solar Spectrum

We tested the results from the two-stream code against a multi-stream discrete ordinate code (DISORT – Stammes et al., 1988) for the case of gaseous absorption and molecular scattering in the solar spectrum. The gaseous absorption coefficients were identical for both the two-stream and the DISORT sixteen-stream calculations. The results for the downward flux at the surface, upward flux at the top of the atmosphere and absorbed flux (in W/m^2) are shown in Table 6.

As shown, the differences between the two-stream and DISORT sixteen-stream results are less than one W/m^2 . This is not particularly surprising since Rayleigh scattering is relatively isotropic and most of the energy is in the direct beam. The ICRCCM column is the result from Fouquart et al., (1991) as an average of a number of radiative transfer models. It shows the spread for 30 models for this relatively simple case.

The band by band comparison is shown in Figure 1. The figure shows that the majority of the energy is in bands 3-6 (wavelength region $0.345\text{-}1.24 \mu\text{m}$). The absorption by ozone in the UV and by water vapor in the near infrared is also apparent.

3.3 Case 3 - Aerosol scattering

Since the model is designed for use with aerosol and cloud models, we need to know the accuracy for the case of aerosol and cloud absorption and scattering. However, there are a myriad of cases that can be run (see King and Harshvardhan, 1986 for examples) and we can not present results for all possible cases. We will only present some representative aerosol results here.

3.3.1 Solar Spectrum

The largest error in the two stream scattering approximation is for aerosol (and cloud) scattering in the solar spectrum. Toon et al., (1989) show that the approximation should be accurate to within 10% in the diffuse (scattered) flux. For clear sky this means that the method should be roughly 1% in the total flux (direct + diffuse) as discussed by Kato et al., (1997). However, this one percent can mean up to 10 W/m².

We calculated the solar flux for a base case of Midlatitude Summer, solar zenith angle 30 deg, surface albedo of 0.2 (ICRCCM-Case 31, Foquart et al., 1991). This case is representative of much of the ARM clear sky data base (Kato et al., 1998) and was kept constant for simplicity.

We used two different aerosols for illustration: a mineral aerosol and a continental/urban aerosol. The mineral aerosol chosen was the WCP-55 (WMO 1983) mineral aerosol as it has a low single scattering albedo (large absorption) and a relatively flat spectral variation compared to the other mineral aerosol data (Sokolik and Toon, 1996). The continental/ urban aerosol was that with properties observed in the recent TARFOX experiment conducted over the eastern U.S. in 1996 (Russell, et al., 1998). The radiative properties used are shown in Table 7.

The downward solar flux at the surface and the upward solar flux as a function of aerosol optical depth are shown in Figure 2 for both the two- stream and DISORT sixteen-stream calculations for both the TARFOX and WCP-55 mineral aerosol. As shown, the decrease in the solar flux at the surface is considerable, with the WCP-55 mineral aerosol significantly reducing the downward flux. For the WCP-55 mineral aerosol, the upward flux at the top of the atmosphere decreases, indicating a warming effect (See Sokolik and Toon, 1996 - Figure 3). However, the TARFOX aerosol shows relatively little effect on the upward flux, thus, neither heating nor cooling the earth-atmosphere system. (The TARFOX aerosol actually showed a cooling effect over the ocean, due to a lower surface albedo, Hignett et al., (1998).)

Figure 2 shows that the error caused by use of the two-stream method is considerably less than the difference between the two aerosol models. The error in using only a few streams as compared with

more streams is shown in Figure 3 for the WCP-55 aerosol case. The figure shows the fall off of the error with increasing number of streams for the flux at the surface.

The effect of neglecting the absorption is shown in Figure 4. The amount of aerosol absorption is an extremely difficult measurement (Bohren and Huffman, 1983; Horvath, 1993; Sokolik and Golitsyn, 1993) and often is simply unknown for many atmospheric measurement programs. As shown in Figure 4, neglect of the absorption (if present) can result in considerable error. Again, for these two cases, the neglect of the aerosol absorption would lead to more error than the use of a two-stream scattering code.

3.3.2 Thermal Spectrum

Toon et al., (1989) evaluated the two stream approximation for the emissivity of an isothermal layer for several single scattering albedos. They showed that source function method was superior to the other two stream approximations and that the maximum error in the emissivity was about 10%. Sokolik et al., (1998) have recently used this model to compute the thermal radiation forcing from mineral aerosols.

4. Additional discussion

We envision the primary use of this radiative transfer model to be inside climate models that calculate the radiative effects of clouds and aerosols together with other effects, such as droplet growth, size distribution change, etc (refs). There are several issues that warrant some discussion:

- Spectral resolution for the clouds and aerosols

A frequently used simplifying assumption is to ignore the spectral dependence of the radiative properties of the cloud or aerosol.

- Computational efficiency

The speed of a computer model depends on several factors. This model has a total of 480 spectral computation intervals. We believe that the number could be reduced in certain cases.

The two stream method has the beneficial property of being an explicit calculation. We have experimented with a four stream approach, but have not been able to reduce the amount of time to less significantly less than N2 (factor of 16).

5. Summary and Conclusion

We have developed an improved radiative transfer model for use in climate models. We have compared the results in the solar spectrum to more detailed calculation schemes. The results indicate that the model is accurate within 1 W/m² in transmission through the atmospheric gases and molecular scattering. For aerosol scattering the two stream code overpredicts the radiant flux at the surface by about 1 %. This difference is less than the current uncertainty between measurements and predictions of the solar radiant flux and less than the effect of not knowing the aerosol radiative properties.

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Table 1 The Spectral Bands

Thermal Bands						
Band	Wavenumber Range (cm-1)	Lower Atmosphere		Upper Atmosphere		Wavelength Range (μm)
		Key Species	Minor Species	Key Species	Minor Species	
1	10-250	H ₂ O		H ₂ O		40-100
2	250-500	H ₂ O		H ₂ O		20-40
3	500-630	H ₂ O, CO ₂		H ₂ O, CO ₂		15.87-20
4	620-700	H ₂ O, CO ₂		CO ₂ , O ₃		14.29-15.87
5	700-820	H ₂ O, CO ₂	CCL ₄	CO ₂ , O ₃	CCL ₄	12.20-14.29
6	820-980	H ₂ O	CO ₂ , CFC-11/12	-----	CO ₂ , CFC-11/12	10.20-12.20
7	980-1080	H ₂ O, O ₃ , CO ₂		O ₃		9.26-10.20
8	1080-1180	H ₂ O	CO ₂ , CFC-12/22	O ₃	CO ₂ , CFC-12/22	8.47-9.26
9	1180-1390	H ₂ O, CH ₄		CH ₄		7.19-8.47
10	1390-1480	H ₂ O		H ₂ O		6.76-7.19
11	1480-1800	H ₂ O		H ₂ O		5.56-6.76
12	1800-2080	H ₂ O, CO ₂		-----		4.81-5.56
13	2080-2250	H ₂ O, N ₂ O		-----		4.44-4.81
14	2250-2380	CO ₂		CO ₂		4.20-4.44
15	2380-2600	N ₂ O, CO ₂		-----		3.85-4.20
16	2600-3250	H ₂ O, CH ₄		-----		3.33-3.85

Solar Bands						
Band	Wavenumber Range (cm-1)	Lower Atmosphere		Upper Atmosphere		Wavelength Range (μm)
		Key Species	Minor Species	Key Species	Minor Species	
1:	10-2600	H ₂ O		CO ₂		3.846-100
2:	2600-3250	H ₂ O, CH ₄		CH ₄		3.077-3.846
3:	3250-4000	H ₂ O, CO ₂		H ₂ O, CO ₂		2.500-3.077
4:	4000-4650	H ₂ O, CH ₄		CH ₄		2.151-2.500
5:	4650-5150	H ₂ O, CO ₂		CO ₂		1.942-2.151
6:	5150-6150	H ₂ O;	CH ₄	H ₂ O		1.626-1.942
7:	6150-7700	H ₂ O, CO ₂		H ₂ O, CO ₂		1.299-1.626
8:	7700-8050	H ₂ O, O ₂		O ₂		1.242-1.299
9:	8050-12850	H ₂ O;		-----		0.778-1.242
10:	12850-16000	H ₂ O, O ₂	O ₃	O ₂	O ₃	0.625-0.778
11:	16000-22650	H ₂ O;	O ₃	-----	O ₃	0.442-0.625
12:	22650-29000	-----		-----		0.345-0.442
13:	29000-38000	O ₃ ;		O ₃		0.263-0.345
14:	38000-50000	O ₃ , O ₂		O ₃ , O ₂		0.200-0.263

Table 2. Range of aerosol values of single scattering albedo, asymmetry factor and optical depth for the solar spectrum (a) and the thermal spectrum (b). The values are illustrative only and not meant to be a definitive compilation. The references are (1) Sokolik and Toon, 1996, (2) D'Alimena et al 1991, (3) Sokolik et al, 1998, (4) Horvath, 1993, (5) Toon and Pollack, 1977

<u>Aerosol component</u>	<u>ω</u>	<u>g</u>	<u>τ</u>	<u>Reference</u>
Mineral Dust	0.65 - 0.95	0.65 - 0.88	0.10 - 1.00	(1)
Sulfate*	0.92 - 1.00	0.50 - 0.80	0.01 - 0.50	(2)
Black Carbon	0.10 - 0.30	0.10 - 0.40	0.01 - 0.10	(4)
Volcanic	0.99 - 1.00	0.50 - 0.70	0.01 - 0.10	(5)

(a) : Solar spectrum 0.2-3.0 μm ; 3,333 – 50,000 cm^{-1}

* The component “Sulfate” represents the non-absorbing and weakly absorbing secondary aerosols including carbonaceous material.

<u>Aerosol component</u>	<u>ω</u>	<u>g</u>	<u>τ</u>	<u>Reference</u>
Mineral Dust	0.40 - 0.88	0.50 - 0.80	0.10 - 1.00	(3)
Sulfate	0.00 - 0.20	0.00 - 0.40	0.01 - 0.50	(2)
Volcanic	0.20 - 0.90	0.50 - 0.70	0.01 - 0.10	(5)

(b) : Thermal spectrum 10-3,333 cm^{-1} (3.0 – 100 μm)

Table 3. Radiative Flux at surface for the k distribution method and the line by line code LBLRTM.

<u>Band</u>	<u>Wavenumber, cm⁻¹</u>	<u>TOA</u>	<u>k-distribution</u>	<u>LBLRTM</u>
1	10-2600	13.2	3.84	3.77
2	2600-3250	12.1	6.33	5.85
3	3250-4000	20.4	0.33	0.46
4	4000-4650	23.7	17.87	17.73
5	4650-5150	22.4	14.42	14.15
6	5150-6150	55.6	29.08	29.68
7	6150-7700	102.9	47.33	47.62
8	7700-8050	24.3	23.52	23.48
9	8050-12850	345.7	281.73	282.42
10	12850-16000	218.2	203.26	203.02
11	16000-22650	347.2	338.58	338.43
12	22650-29000	129.5	129.46	129.46
13	29000-38000	48.4	26.70	27.39
14	38000-50000	3.1	0.00	0.01
TOTAL		1366.7	1122.45	1123.47
Absorbed flux =			244.25	243.23

Table 4. Radiative fluxes for the case of Rayleigh Scattering and Gaseous absorption.

a. Downward flux at the surface (W/m²)

<u>Atmosphere</u>	<u>μ_0</u>	<u>2 Stream</u>	<u>16 streams</u>	<u>ICCRM</u>
MLS				
	.866	921.56	921.25	943.7 (+/- 25)
	.2588	225.14	224.47	235.8
Tropical				
	.866	907.48	907.17	932.6
	.2588	221.56	220.85	234.9

b. Upward flux at the top of the atmosphere (W/m²)

<u>Atmosphere</u>	<u>μ_0</u>	<u>2 Stream</u>	<u>16 streams</u>
MLS			
	.866	210.23	209.21
	.2588	77.71	77.16
Tropical			
	.866	208.41	207.56
	.2588	77.72	77.26

c. Absorbed flux in the atmosphere (W/m²)

<u>Atmosphere</u>	<u>μ_0</u>	<u>2 Stream</u>	<u>16 streams</u>	<u>ICCRM</u>
MLS				
	.866	235.80	237.13	206.2 (+/-12)
	.2588	95.78	96.89	83.8
Tropical				
	.866	248.90	250.04	215.1
	.2588	98.63	99.64	84.0

TABLE 5. The spectral radiative properties of the WCP-55 and TARFOX average aerosol

Band	WCP-55			TARFOX		
	$\underline{\omega}$	g	$\tau/\tau_{.5}$	$\underline{\omega}$	g	$\tau/\tau_{.5}$
1	0.54	0.95	0.97	0.92	0.77	1.86
2	0.58	0.92	0.97	0.91	0.76	1.50
3	0.63	0.89	0.99	0.91	0.73	1.29
4	0.66	0.88	1.00	0.90	0.70	1.00
5	0.68	0.86	1.01	0.88	0.66	0.57
6	0.71	0.84	1.04	0.85	0.62	0.29
7	0.73	0.83	1.06	0.84	0.62	0.29
8	0.75	0.83	1.07	0.83	0.68	0.14
9	0.78	0.86	1.07	0.83	0.69	0.13
10	0.79	0.89	1.03	0.83	0.70	0.13
11	0.80	0.90	1.00	0.83	0.72	0.14
12	0.80	0.90	1.00	0.50	0.75	0.14
13	0.80	0.90	1.00	0.30	0.77	0.14
14	0.80	0.90	1.00	0.20	0.78	0.14

List of Figures

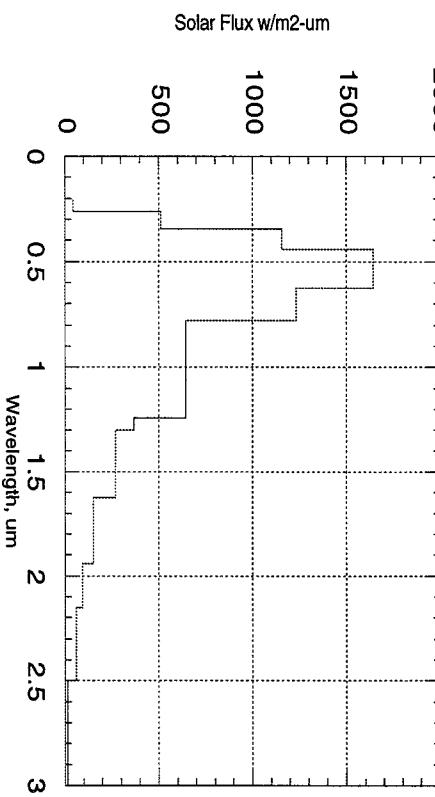
Figure 1: Spectral distribution of the solar flux for the case of Rayleigh scattering and gaseous absorption

Figure 2: Solar Flux as a function of aerosol optical depth for different aerosol models and multi-stream methods

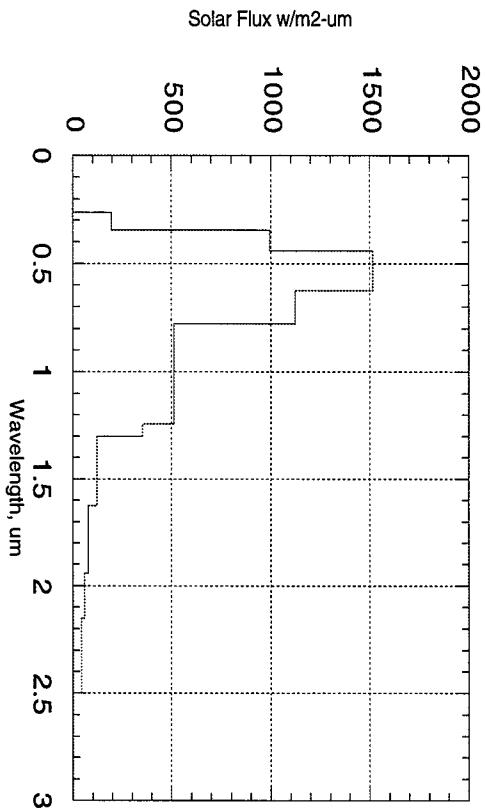
Figure 3: Error in downward flux at surface for using various multi-stream methods

Figure 4: Solar Flux as a function of optical depth with and without aerosol absorption

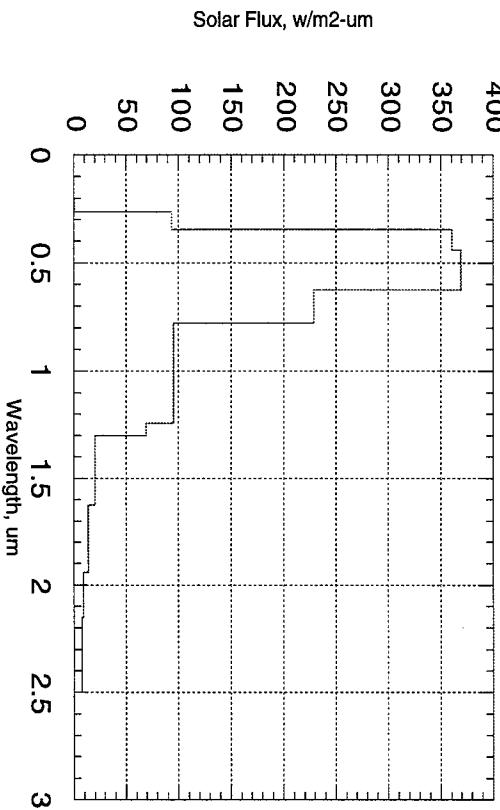
Downward Flux at the top of atmosphere

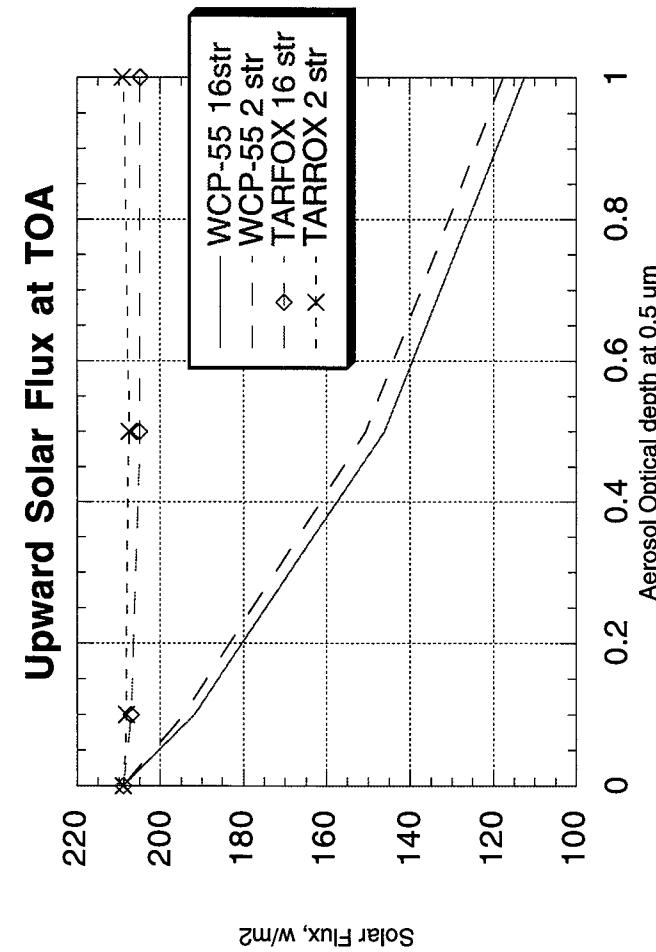
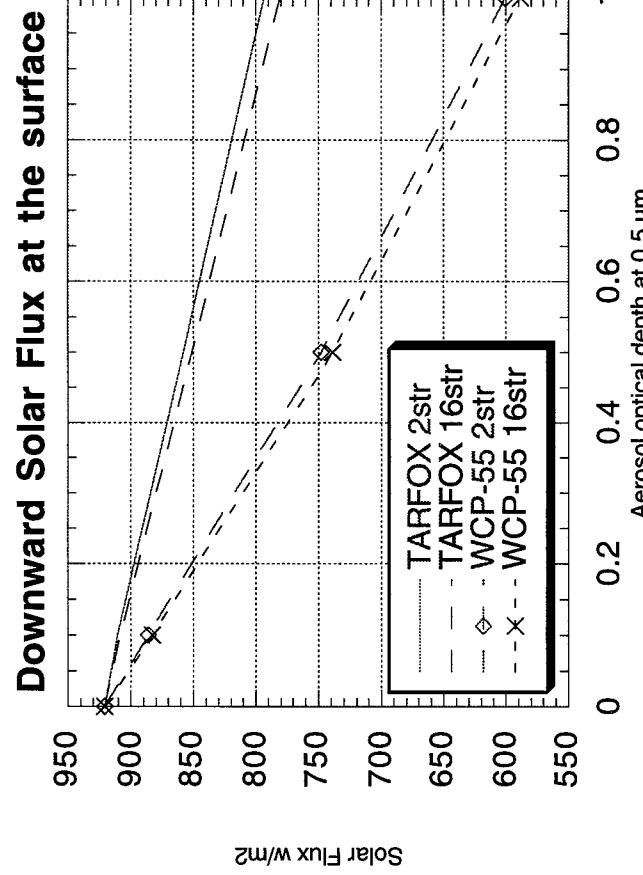


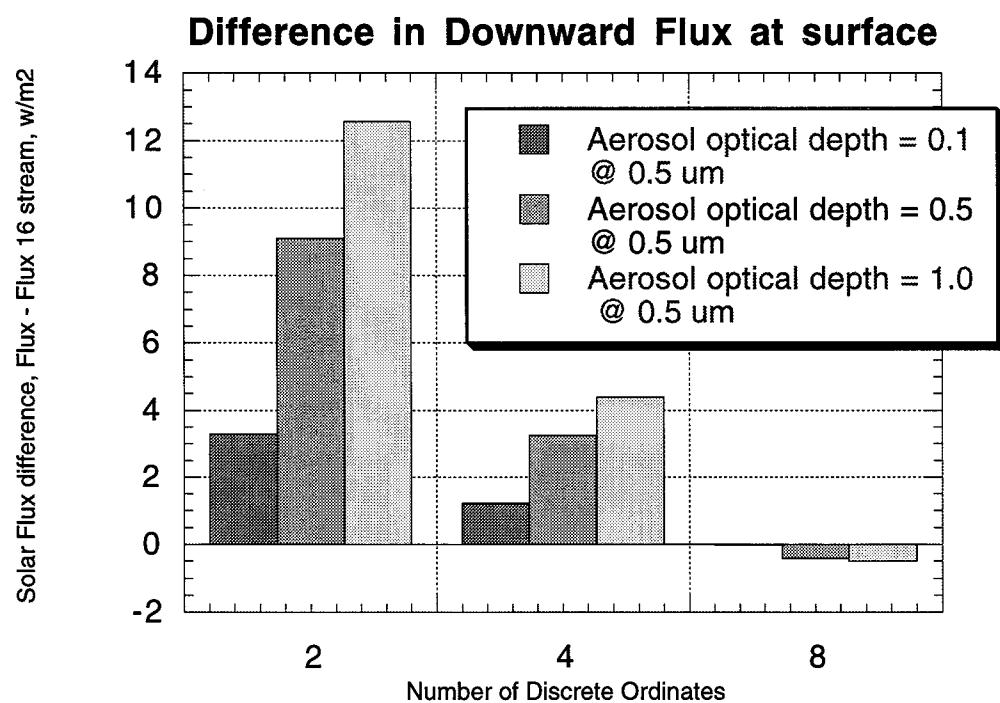
Downward Flux at the Surface



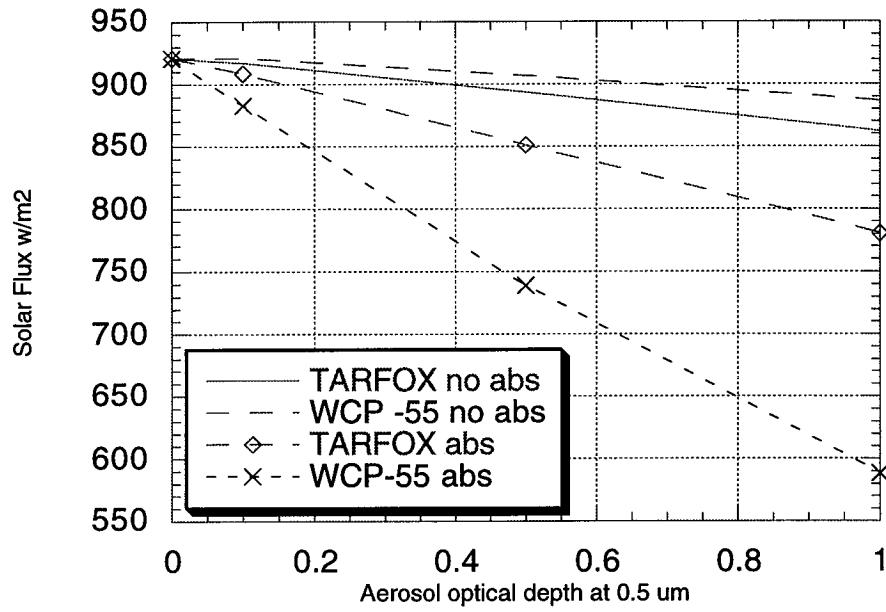
Upward Flux at the top of the atmosphere







Downward Flux at the surface



Upward Flux at TOA

